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SHORT NOTES

Ductility of garnet as an indicator of extremely high temperature deformation: Discussion

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In a paper entitled "Ductility of garnet as an indicator of extremely high temperature deformation", Ji & Martignole (1994) argue that garnets were flattened substantially by dislocation creep in the Morin shear zone (Grenville Province, Quebec, Canada) and that the temperature during deformation must therefore have exceeded 900°C. Ji & Martignole (1994) further argue that quartz is stronger than garnet under these high temperature conditions. We are not convinced by their arguments. The deformation of the garnet studied by Ji & Martignole (1994) is much better explained in terms of grain boundary diffusional creep (e.g. Coble creep, or pressure solution) at a temperature in the range 550 to 750° C, as was previously estimated to be the temperature when the Morin shear zone was active (Martignole & Schrijver 1970, Martignole 1992).

Ji & Martignole (1994) base their point that the garnets deformed by dislocation creep on a transmission electron microscopy (TEM) study of the deformed garnets. They observed (see their fig. 5): (i) a quite homogeneous distribution of free dislocations, (ii) the presence of both curved and straight dislocations, (iii) subgrain boundaries composed of dislocation arrays, and (iv) locally extensively developed dislocation networks. For Ji & Martignole (1994) these observations indicate that the garnet deformed by steady-state creep due to "effective dynamic recovery (dislocation climb and cross-slip)". According to them, recovery-accommodated dislocation creep was responsible for the entire deformation of the garnet. We think, however, that these TEM observations alone are not exclusive enough to argue that the garnet must have been deformed entirely by dislocation creep. The observations are consistent with, but by no means exclusive proof of, significant amounts of strain by dislocation processes. The total amount of strain by dislocation motions might have been very small.

According to Ji & Martignole (1994) the garnets were flattened by about 50% on average, almost entirely by dislocation creep. Yet, the garnets show no evidence of recrystallisation. Ji & Martignole (1994) argue, on the basis of their TEM work, that dislocation glide was made possible by dynamic recovery. So, one would expect that subgrain rotation was important. In the case of quartz or feldspar, much lower strains than 50% are enough for recrystallisation by progressive subgrain rotation. For example, strains of less than 50% lead to intense recrystallisation of garnets (Fig. 1) in high grade metamorphic rocks that experienced temperatures between ≈700 and 800° C (Schenk 1990).

A flattening of the order of 50% by dislocation creep should lead to a significant crystallographic preferred orientation (CPO). The presence of a CPO would support the point of Ji & Martignole (1994). However, whether or not a CPO was present, was not investigated. This was done recently by Kleinschrodt & McGrew (1995) for morphologically similar garnets from granulite facies rocks that were deformed at a temperature in the range 800 to 850° C and a pressure of the order of 8 to 9 kbar. An SEM electron channeling pattern analysis showed that the garnets had a random crystallographic orientation. Kleinschrodt & McGrew (1995) therefore argued that dislocation creep could not have been important, and that diffusional creep had possibly been the dominant deformation mechanism in the rocks they studied.

Ji & Martignole (1994) report that some of the garnets contain relatively large quartz inclusions. These inclusions are spherical, and show no optical evidence for intracrystalline deformation. An example is shown in fig. 4(e) of Ji & Martignole (1994); compare with Fig. 2. The absence of internal strain features in the quartz inclusions, as well as their spherical shape, would indicate that the garnets remained undeformed internally, and that their change in shape must be due to removal of material at grain boundaries by grain boundary diffusional deformation processes (e.g. by Coble creep, or pressure solution). Ji & Martignole (1994) suggest, however, that the quartz inclusions remained undeformed because quartz was much stronger than the garnet under the high *P–T* conditions at which the garnets were deformed. According to them, deformation took first place under relatively high *P–T* conditions, at $T > 900°$ C, and then under relatively low *P–T* conditions. At high *P–T* conditions, the garnets must have been soft inclusions in a relatively stiff, stress supporting, quartzo–feldspathic matrix. At low *P–T* conditions, the garnets must have been much stronger than the quartzo–feldspathic matrix, and consequently have protected the quartz inclusions from deformation. Following this scenario, the garnet must have flowed around the quartz inclusions at the high *P–T* conditions and crystal plastic deformation by glide and climb of dislocations must have been very heterogeneously distributed. Yet, there are no remarkable optical signs of heterogeneous strain (deformation bands, recrystallisation features, fractures around the inclusions), not even in those regions where the quartz inclusions approach the garnet rim as near as $\approx 10\ \mu$m (see fig. 4(e) of Ji & Martignole 1994 and compare with Fig. 2). The rim of the garnet shows no sign of bending around the quartz inclusion in Fig. 2.

Rigid feldspars included in a relatively soft and ductile quartz rich matrix are often fractured, with quartz occurring as fracture infill. Ji & Martignole (1994) make no mention of fractured quartz inclusions, with garnet as fracture infill.

One argument that Ji & Martignole (1994) put forward in support of garnet being weaker than quartz and feldspar in the matrix under high *P–T* conditions, is that the garnets are oblate ($K = 0.126$) whereas the quartz and feldspar are prolate ($K = 1.02$). They refer to the work of Freeman & Lisle (1987) who have shown theoretically that during a single deformation event, material stiffer than the matrix will deform into a less oblate shape than the matrix itself, and vice versa. Following Ji & Martignole (1994) the garnets are oblate because they were weaker, the quartz and feldspar in the matrix prolate because they were stronger. However, Ji & Martignole (1994) conclude themselves that the rocks in the Morin shear zone underwent two deformation events: "Ductile strain recorded by the garnet grains first took place at extremely high temperature and was then preserved from further intracrystalline deformation upon cooling down to metamorphic temperatures. It was followed by a considerably large, rotational strain ($\gamma = 10$, Martignole 1992) which took place at temperatures straddling the granulite–amphibolite facies conditions". The problem is that the possibility cannot be excluded, that the prolateness of the quartz and feldspar in the matrix was developed in the second event at the lower temperature, and that the quartz and feldspar originally deformed into an oblate shape during the same event as the garnet was deformed in such a shape. In this case, the oblateness of the garnet cannot be used as an argument for it having been weaker than quartz and feldspar.

In the case of garnet, it may be correct to assume an equidimensional shape prior to deformation, but not for feldspar and quartz. Their shape and arrangement may have been influenced by numerous processes prior to and during deformation (e.g. by solid state or fluid assisted diffusional deformation processes, by metamorphic reactions, by stress or strain driven grain boundary migration). Without knowledge of the original grain shape, a strain measurement cannot be made.

By extrapolating experimentally determined flow laws for garnet and for quartz to natural conditions, Ji & Martignole (1994) try to demonstrate that garnet must indeed be weaker than quartz at a temperature above 900° C and a strain rate of the order of $10^{-14}\ s^{-1}$ (see their fig. 8). For garnet, Ji & Martignole (1994) used unpublished parameters obtained by personal communication from Wang and Karato in experiments on almandine–pyrope (for the parameters see Table 1). They also refer to an EOS-abstract (Liu *et al.* 1992), but the parameters used by Ji & Martignole (1994) cannot be found there. In fact, the exact experimental conditions at which the parameters were obtained cannot be traced back.

For the flow behaviour of quartzite, Ji & Martignole (1994) used the parameters from the EOS-abstract of Koch *et al.* (1980). These parameters are out of date.

Table 1

	A [$MPa^{-n}\ s^{-1}$]	n	Q [kJ/mole]
Quartzite: Jaoul *et al.* (1984) vac. heated	$3.44 \cdot 10^{-6}$	2.8 ± 0.2	185 ± 6
Quartzite: Jaoul *et al.* (1984) as is	$9.98 \cdot 10^{-6}$	2.4 ± 0.4	163 ± 13
Quartzite: Jaoul *et al.* (1984) wet	$5.26 \cdot 10^{-3}$	1.4 ± 0.05	146 ± 4
Quartzite: Koch *et al.* (1989) dry	$1.16 \cdot 10^{-7}$	2.72 ± 0.19	134 ± 32
Quartzite: Koch *et al.* (1989) wet	$5.05 \pm 5 \cdot 10^{-6}$	2.61 ± 0.15	145 ± 17
Quartzite: Koch *et al.* (1980) wet	$2.4 \cdot 10^{-7}$	2.86 ± 0.18	149 ± 29
Garnet: Wang & Karato (pers. comm.) referred to by Ji & Martignole (1994)	$5.84 \cdot 10^{6}$	2.22 ± 0.1	485 ± 30

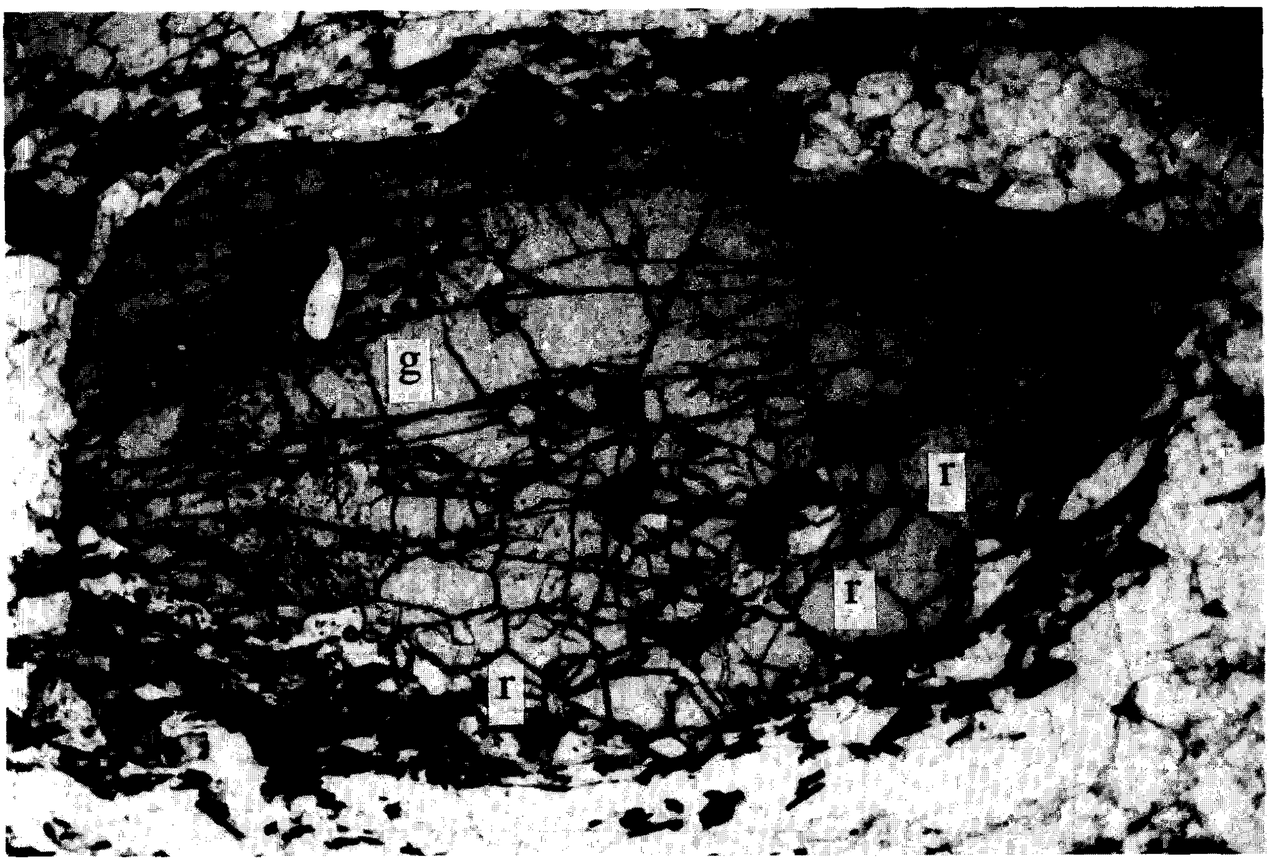

Fig. 1. Granulite facies garnet porphyroblast from the central metapelite unit of the Calabrian lower crustal section (Schenk 1990). Polygonal, inclusion-free recrystallised grains (some of which are marked 'r') are developed along the lower right margin of the blast (marked 'g'). The blast itself contains inclusions of fibrolitic sillimanite, quartz, and graphite. (Sample number Kr3390; section perpendicular to the foliation; parallel polarisers; long side of photograph is 3.5 mm.)

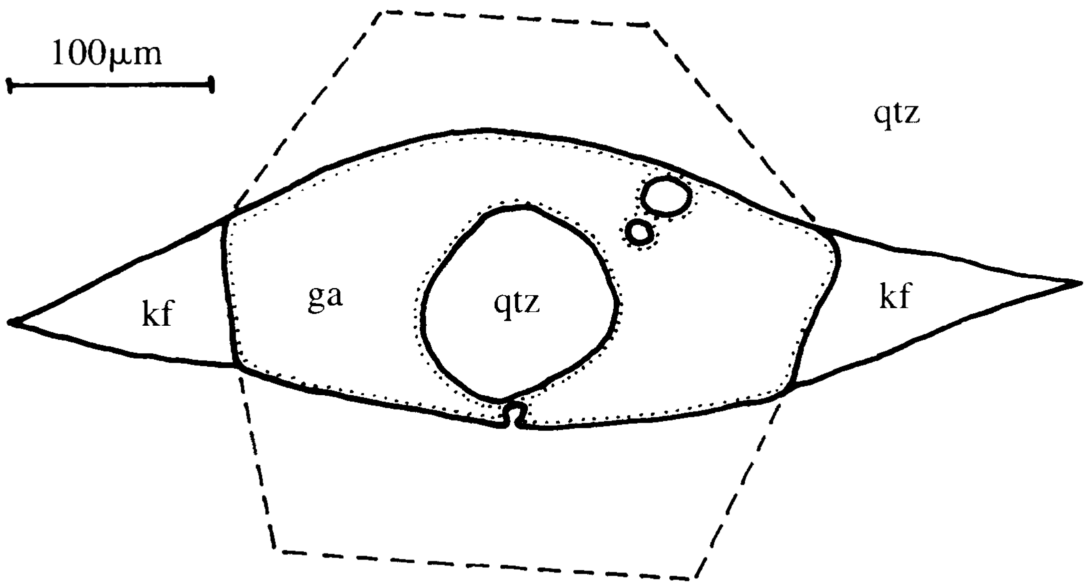

Fig. 2. Schematic drawing of the deformed lenticular garnet single crystal in the quartzo-feldspathic mylonite such as presented in fig. 4(e) of Ji & Martignole (1994). *X–Z* section. ga, garnet; kf, K-feldspar; qtz, quartz.

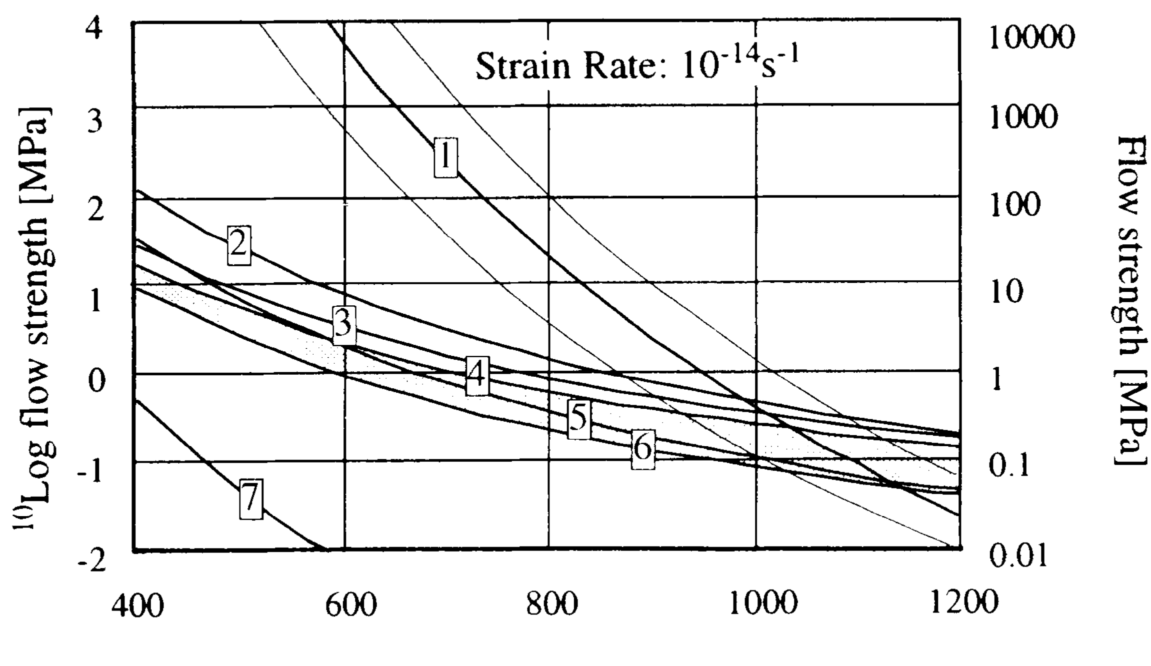

Fig. 3. Flow strength versus temperature profiles for experimentally determined crystal plastic deformation processes. (1) Garnet (Alm68 Prp20 Grs12) after Wang & Karato (pers. comm. referred to by Ji & Martignole 1994). Thin lines on either side of line 1 indicate upper and lower error range. (2) Vacuum dried quartzite (Jaoul *et al.* 1984). (3) Quartzite in presence of dehydrating talc confining medium (Koch *et al.* 1980). (4) As-is quartzite (Koch *et al.* 1989). (5) As-is quartzite (Jaoul *et al.* 1984). (6) Quartzite in presence of dehydrating talc confining medium (Koch *et al.* 1989). (7) Quartzite deformed with ~0.4 wt% added water (Jaoul *et al.* 1984). Field between lines 4 and 6 is shaded.

Significantly lower flow stresses are predicted by more recent work of, e.g. Jaoul *et al.* (1984) and Koch *et al.* (1989) (see Table 1 and Fig. 3). The temperature at which the garnet line of Ji & Martignole (1994) crosses the quartzite lines is consequently higher than the 900° C suggested by Ji & Martignole (1994). It lies at approximately $1000 \pm 100°$ C for completely dry quartzite (see line 2 in Fig. 3 for natural quartzite vacuum heated at 800° C for 12 hr) and at about $1100 \pm 100°$ C for quartzite containing up to about 0.1 wt% (intra- or intercrystalline) water (lines 4 and 5 in Fig. 3 for as-is deformed natural quartzite). Note, that these temperatures are at or even above the melting point of the completely dry Albite–Orthoclase–Quartz system ($Tm = 950°$ C at $P = 500$ MPa and $Tm = 1000°$ C at $P = 1000$ MPa; see, e.g. fig. 18–11 in Winkler 1979).

The steady state flow stresses predicted by the extrapolation of the experimentally determined flow laws are unrealistically low (0.01–1 MPa; see Fig. 2 or fig. 8 of Ji & Martignole 1994). Ji & Martignole (1994) do not discuss this point. At such low stresses, the glide process might very well be the rate controlling process, rather than the climb process. If so, then the garnet should always be stronger than the quartz because it has a higher Peierls stress.

According to us, Ji & Martignole (1994) fail to demonstrate that the garnets were deformed substantially by dislocation creep. The lens-shape of garnet is much better explained in terms of grain boundary diffusional creep (e.g. Coble creep or pressure solution), as is strongly suggested by the microphotographs presented by Ji & Martignole (1994); see especially their fig. 4(e). A dominance of grain boundary diffusional creep, with only minor dislocation creep, would be consistent with the TEM observations. It would be the simplest explanation for the occurrence of undeformed quartz inclusions in the garnets, and the conspicuous absence of optical lattice strain features in the garnets, such as the absence of recrystallised grains. Furthermore, garnet need not be weaker than quartz, and the temperature need not be extremely high. A dominance of grain boundary diffusional creep of garnets would be in good agreement with estimates of Martignole & Schrijver (1970) and Martignole (1992) that the Morin shear zone was active at temperatures in the range 750 to 550° C (decreasing during its activity). As to the TEM observations of Ji & Martignole (1994) (i.e. the dislocation substructures held indicative of dynamic recovery), it can be deduced from Fig. 3, that at geologically realistic differential stresses in the range 10 to 1000 MPa, and temperatures in the range 550 to 750° C, garnet can be deformed by dislocation creep at geologically realistic strain rates. Hence, assuming that the extrapolation of the type done in Fig. 3 is valid, it is very well possible that during activity of the Morin shear zone at 550 to 750° C, some crystal plastic deformation of garnets took place, but at significantly lower deformation rates than in quartz and feldspar.

REFERENCES

Freeman, B. & Lisle, R. J. 1987. The relationship between tectonic strain and the three-dimensional shape fabrics of pebbles in deformed conglomerates. *J. geol. Soc. Lond.* **144**, 635–639.

Jaoul, O., Tullis, J. & Kronenberg, A. 1984. The effect of varying water contents on the creep behaviour of Heavitree quartzite. *J. geophys. Res.* **89**, 4298–4312.

Ji, S. & Martignole, J. 1994. Ductility of garnet as an indicator of extremely high temperature deformation. *J. Struct. Geol.* **16**, 985–996.

Kleinschrodt, R. & McGrew, A. J. 1995. Garnet plasticity in the lower continental crust: Constraints on deformation mechanisms from microstructural and textural data. *J. Czech. geol. Soc.* **40**, 104.

Koch, P. S., Christie, J. M. & George, R. P. 1980. Flow law of wet quartzite in the alpha quartz field. *Eos* **61**, 376.

Koch, P. S., Christie, J. M., Ord, A. & George, R. P. 1989. Effect of water on the rheology of experimentally deformed quartzite. *J. geophys. Res.* **94**, 13975–13996.

Liu, B., Wang, Z., Fujino, K. & Karato, S. 1992. High-temperature creep of garnets, III: Alumino–silicate garnets and the rheology of the transition zone. *Eos* **73**, 378.

Martignole, J. 1992. The Morin terrane revised. In: *Abstract for Abitibi–Grenville* Lithroprobe Workshop (Montréal, Canada), 15.

Martignole, J. & Schrijver, K. 1970. Tectonic setting and evolution of the Morin Anorthosite Grenville Province, Quebec. *Bull. geol. Soc. Finland* **42**, 165–209.

Schenk, V. 1990. The exposed crustal cross-section of southern Calabria, Italy: Structure and evolution of a segment of Hercynian crust. In: *Exposed cross-sections of the continental crust* (edited by Salisbury, M. H. & Fountain, D. M.) pp. 21–42, Kluwer Academic Publishers.

Winkler, H. G. F. 1979 *Petrogenesis of metamorphic rocks.* Fifth Edition, Springer Verlag, New York.